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Ice streaming in the Laurentide Ice Sheet: A first comparison between data-calibrated numerical model output and geological evidence

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[1] Despite the importance of rapidly-flowing ice streams to ice sheet mass balance, their incorporation into numerical ice sheet models is a major scientific challenge. This introduces large uncertainties in model output and inhibits a more complete understanding of the role of ice streams in overall ice sheet stability. Recent computational advances have enabled more realistic representations of ice streaming but few studies have attempted to compare model output against known locations of ice streams. This paper compares predictions of ice streaming derived from a large ensemble analysis of a Glacial Systems Model of the Laurentide Ice Sheet against independent geological evidence compiled from previously published studies. Although the precise dating of paleo-ice stream locations is problematic, our analysis includes comparisons at six different time-steps (18 to 10 cal ka BP) during deglaciation. Results indicate that the model is successful in predicting all of the major marine-terminating ice streams but there is mixed success in simulating terrestrial ice streams in the right place and at the right time, which is vital in guiding future model development. The model also reveals that whilst some ice streams persist throughout deglaciation the focus of mass loss associated with ice streaming switches through time with dynamic changes in ice stream catchments and tributaries. This implies that major changes in ice stream activity are to be expected in a deglaciating ice sheet, with important implications for contemporary ice sheet dynamics. **Citation:** Stokes, C. R., and L. Tarasov (2010), Ice streaming in the Laurentide Ice Sheet: A first comparison between data-calibrated numerical model output and geological evidence, *Geophys. Res. Lett.*, *37*, L01501, doi:10.1029/2009GL040990.

1. Introduction

[2] Mass loss from ice sheets is focused through rapidly-flowing ice streams that account for the majority of ice sheet discharge. These features are highly dynamic and have been shown to accelerate, decelerate, migrate, and stop, over relatively short (decadal) time-scales [e.g., *Anandakrishnan and Alley*, 1997; *Joughin et al.*, 2004; *Conway et al.*, 2002]. Despite their importance, ice streams represent a major uncertainty in our understanding of ice sheet mass balance [*Intergovernmental Panel on Climate Change* (*IPCC*), 2007], partly due to the short time-scale of recent observations (decades), compared to the much longer time-scale of their activity (centuries to millennia). In order to advance our understanding of the processes that initiate, sustain or inhibit ice streaming over time-scales more appropriate to their longevity, many workers have focused on reconstructing paleo-ice stream activity in former ice sheets (see review by *Stokes and Clark* [2001]).

[3] Reconstructing the behavior of paleo-ice streams is usually based on one of two approaches. The first is to reconstruct their dynamics by piecing together evidence from the landforms and sediments left behind by their activity [e.g., *Patterson*, 1998]. The recognition that the subglacial imprint of an ice stream is distinct from non-streaming areas [e.g., *Kleman et al.*, 1997; *Stokes and Clark*, 1999] has resulted in a large population of 'known' ice stream locations from several paleo-ice sheets [e.g., *Winsborrow et al.*, 2004; *Ottesen et al.*, 2005]. The second approach utilizes numerical models of paleo-ice sheets that are able to simulate ice streaming [e.g., *Boulton and Hagdorn*, 2006]. Thermo-mechanical feedbacks in topographic troughs and the simple parameterization of ice-bed coupling strength over soft sediments generate streaming flow [see *Marshall et al.*, 1996] but it is more difficult to reproduce it over smoother topography and/or hard bedrock. Inter-comparison experiments with simple circular ice sheet models, for example, create different ice stream geometries under the same forcing [*Payne et al.*, 2000], highlighting the need for further model development and the incorporation of horizontal stress gradients, which introduce greater consistency [*Hindmarsh*, 2009].

[4] Given the current attention on improving ice sheet models and constraints on ice dynamics for the next IPCC report, there is an urgent need to understand the properties (i.e., boundary conditions) that initiate, sustain or inhibit ice streaming. This can be achieved through a combined approach that evaluates model predictions of ice streaming against independent geological evidence of their activity but very few studies have attempted such a comparison for an entire ice sheet. In this paper, we build on recent advances in data-calibrated numerical modeling and paleo-ice stream identification and report the first data-model comparison of streaming flow in the Laurentide Ice Sheet (LIS), with the aim of providing an improved understanding of the fundamental controls on ice stream activity and their evolution during deglaciation.

2. Methods

[5] It has long been recognised that ice streams are required to produce reconstructions of the LIS that comply

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with geological evidence of ice surface profiles [e.g., *Mathews*, 1974; *Fisher et al.*, 1985]. Early attempts at reconstructing their location [e.g., *Dyke and Prest*, 1987] were augmented with pioneering studies that identified specific geological evidence of their activity, such as distinctive erratic dispersal trains with abrupt lateral margins [*Hicock*, 1988; *Dyke and Morris*, 1988]. Based on these studies, several 'geomorphological criteria' were proposed to aid the identification of paleo-ice streams [*Stokes and Clark*, 1999], including those in the LIS that were compiled by *Patterson* [1998] and more recently inventoried by *Winsborrow et al.* [2004]. We utilize and update this inventory and compare it with the predicted location of ice streams generated by a calibrated Glacial Systems Model (GSM).

[6] The Memorial University of Newfoundland/University of Toronto GSM [*Tarasov and Peltier* [2004, 2005] incorporates a 3D thermomechanically-coupled shallow ice model; a bed thermal module; a visco-elastic bedrock response model using the VM2 earth rheology [*Peltier and Jiang*, 1996]; a fully coupled downslope (diagnostic) surface drainage and surface/proglacial storage (lake) solver; thermodynamic lake ice; a detailed temperature-dependent degree-day mass-balance module; buoyancy and temperature-dependent marine ice calving; thermodynamically constrained lacustrine calving (i.e., taking into account, approximately, the melt potential of the proglacial lake); and shut-down of regional calving when all relevant drainage outlets have less than 20 m minimum water depth. Ice shelves are represented by fast linear basal flow. Streaming in the model requires warm-based ice and, as implemented, occurs weakly via sliding on hard bed, or much more strongly in the presence of sub-glacial till. Relative sea level is computed gravitationally self-consistently except for a eustatic approximation for load corrections due to marine/terrestrial transitions [*Tarasov and Peltier*, 2004]. There is no explicit subglacial hydrology in the current version of the model.

[7] Major sources of uncertainty are assigned parameters subject to ongoing Bayesian calibration through a combination of large model ensembles, artificial neural networks, and Markov Chain Monte Carlo techniques. This involves the calibration of 31 ensemble model parameters (5 are ice-dynamical, 4 for ice calving, 2 for margin forcing, and the remaining 20 associated with the climate forcing) against a large and diverse set of observational constraints. The latter includes a large set of relative sea level (RSL) observations; the best available reconstruction for the ice margin chronology [*Dyke*, 2004], and observed present-day rates of surface uplift (RR). Strandline (paleo-lake level indicator) data and marine limit observations are further used to score model results and therefore determine weighted ensemble results [*Tarasov and Peltier*, 2005]. The location of fast ice flow is not prescribed but evolves freely when the basal ice approaches the pressure melting point, subject to calibrated temporal controls for Heinrich events 1 and 2 and Meltwater Pulse 1a (i.e., two calibration parameters control a temporary increase in regional subglacial till viscosity for a few thousand years prior to these events with return to baseline viscosities at 24 ka (H2), 17.1 ka (H1) and 14.5 ka (Mwp1-a)). Model runs with an Mwpa-1 contribution of <8.5 m are rejected. More complete details on the GSM and constraint data are given by *Tarasov and Peltier* [2004, 2005].

[8] Although model output can be computed at sub-annual time-steps, very few paleo-ice streams have sufficient dating control to enable comparisons at such high temporal resolution. We therefore present model output at 6 time-steps (18, 16, 14, 13, 12, and 10 cal ka BP), which enables valid temporal comparisons for those ice streams which do have some dating control and captures the major spatial changes in ice stream evolution.

[9] Finally, because of the uncertainties associated with past climate and the unavoidable parameterizations required in such models, the comparison of paleo-data with a single model run would have little relevance. We therefore compare the weighted ensemble mean of basal velocity as a proxy for the likelihood of ice streaming. Weighting is with respect to ensemble member fits to the full set of observational constraints. As such, some ice stream locations may be smeared to varying extents.

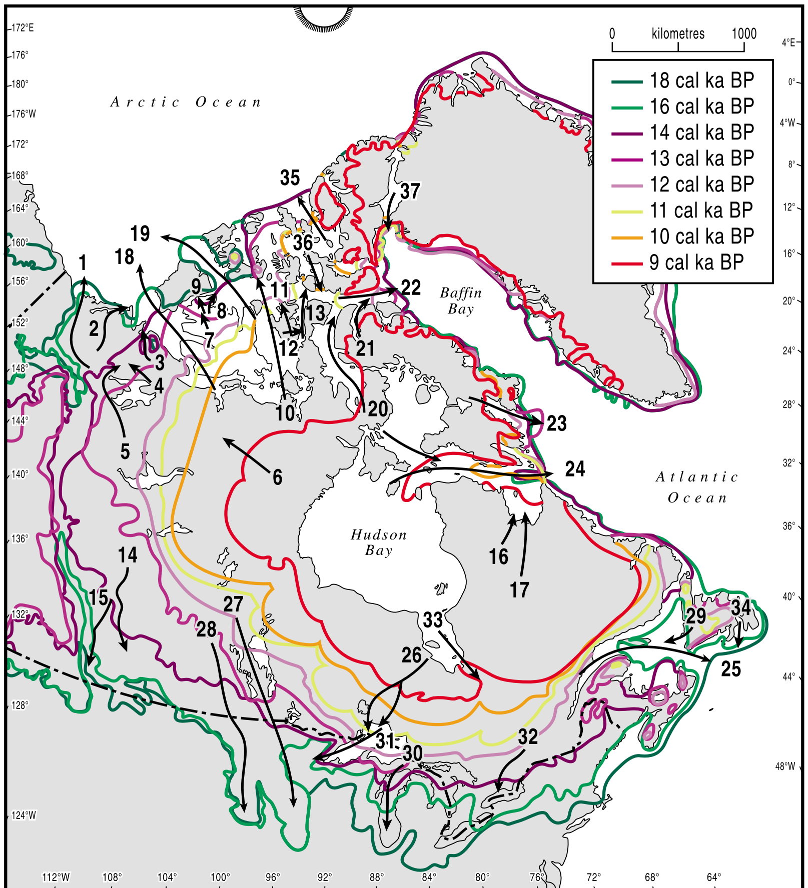
3. Results

[10] The inventory of paleo-ice streams is shown in Figure 1 and the ensemble mean basal velocity of the GSM in Figure 2. Note that model grid-cell size (19–80 km depending on orientation and latitude) precludes small ice streams, which are known to have existed [e.g., *Dyke and Morris*, 1988; *Dredge*, 2000; *De Angelis and Kleman*, 2007] and these will not be discussed further (e.g., 7, 8, 9, and 12 on Figure 1). Our focus in this paper is to provide the first assessment of how well, in general, the model is able to capture major ice streams in the right place and at the right time.

[11] At 18 kyr, four major ice streams terminate in the Arctic Ocean: the Mackenzie (1), Amundsen Gulf (18), M'Clure Strait (19) and Massey Sound (35) ice streams (Figure 1), all of which have been hypothesised based on geological evidence (Table S1 provides ice stream inventory and associated citations).[1] Note areas of high velocity and flat sea-level topography (such as that depicted in the Prince Gustav Adolf Sea between M'Clure Strait and Massey Sound) representing ice shelves. A further three major ice streams drain the eastern margin into the North Atlantic: in Lancaster Sound (22), Hudson Strait (24), and the Laurentian Channel (25). Again, these ice streams match the location of previously identified ice streams (Figure 1). We also note a localised area of rapid velocity that terminates on land at the southern margin of the ice sheet in the vicinity of Lake Michigan/Lake Huron (30). This has also been hypothesised based on geological evidence but there are two further areas of ice streaming located at the southern margin, which have not been previously recognised: one along the border between Alberta and Saskatchewan and one in the Gulf of Maine. Geological evidence for ice streams has been found in Alberta (14 and 15) but they are located further south and east (Figure 1) and their timing is uncertain [*Evans et al.*, 2008]. Elsewhere, localised areas of high velocity (<few grid cells) are modelled in Smith Sound (37) and Cumberland Sound (23), both of which match with previous hypotheses based on glacial geology.

[12] At 16 kyr, major switches are seen in the pattern of ice streams. Marine-based ice streams continue to drain into

[1]Auxiliary materials are available in the HTML. doi:10.1029/2009GL040990.



1 = (21–16)	**8** = 15.2–14.4	**15** = 16.8–15.6	**21** = 18–10	**26** = (18-11)	**32** = 15.6–14.1
2 = (21–15.5)	**9** = 14.4–14.1	**16** = 9.9-8.4	**22** = 18–10	**27** = 16.8 (advance)	**33** = 10–9
3 = (21–14)	**10** = 13–12.65	**17** = 9.9-8.4	**23** = 18–12	**28** = 16.8 (advance)	**34** = 23.5–16
4 = (21–13.5)	**11** = 14.1-13.45	**18** = 21–11.35	**24** = 21–11 (23.2 & 15.5) H-events	**29** = 16.8–16.2	**35** = 21–18
5 = (21–13.5)	**12** = 13.45–9.5	**19** = 21–14.4	**25** = 23.5–16.8	**30** = 18.5–16.5	**36** = 21–18
6 = 10–9	**13** = (21–10)	**20** = 18–10		**31** = (21–13)	**37** = 21–18
7 = ~13	**14** = 16.8–15.6				

All ages cal ka BP

Figure 1. Inventory of paleo-ice streams in the Laurentide Ice Sheet updated from *Winsborrow et al.* [2004] and possible age brackets for their activity. Very few paleo-ice streams are well-constrained by dates but Table S1 provides supporting evidence and citation. Dates in brackets are assigned in this paper based on the location of ice stream 'footprint' in relation to *Dyke*'s [2004] ice margin positions (coloured lines).

the Arctic Ocean but the Lancaster Sound ice stream (22) has lost its major tributary and, in contrast, the Hudson Strait ice stream (24) has enlarged towards the south-west. The terminus of Laurentian Channel ice stream (25) has retreated several hundred kilometres and a new ice stream has switched on in Ungava Bay (17). Ice streams have been hypothesised in the Ungava Bay area but have been estimated to be much younger [*Jansson et al.*, 2003]. Elsewhere the extent of the Mackenzie ice stream (2) appears to be decreasing and subtle switches are seen in the vicinity of Great Bear Lake (5), where complex geological evidence of ice stream flow patterns exist [see *Kleman and Glasser*, 2007]. To the south, a broad zone of high velocity has developed in south-eastern Alberta, where the LIS and Cordilleran ice sheets are 'unzipping'. Geological evidence of ice streams with similar trajectories has not been reported.

[13] The significant change at 14 kyr is the generation of localised ice streams along the southern LIS margin. Geological evidence has been reported from this time, particularly in the Lake Michigan/Superior area (31) and the Lake Winnipeg/Red River are (27, 28). Along the 'unzipping' western margin, the major zone of streaming flow has extended northward and developed an eastward trajectory from the Great Slave Lake. To our knowledge, no geological evidence for this broad zone of streaming flow is reported in the literature but inspection of satellite imagery indicates the presence of localised areas of high elongate bedforms here (at 61°13′ N, 121°42′E), which are often associated with ice streaming [see *Stokes and Clark*, 2001]. Following deglaciation, the Mackenzie (1) and Laurentian Channel (25) ice streams no longer exist; i.e., they have not migrated inland with the ice margin. The Hudson Strait (24) and Lancaster Sound (22) ice streams are similar to their 16 kyr configuration. The termini of the Amundsen Gulf (18) and M'Clure Strait (19) ice streams have retreated several hundred kilometres, but only the M'Clure Strait ice stream has expanded its catchment inland.

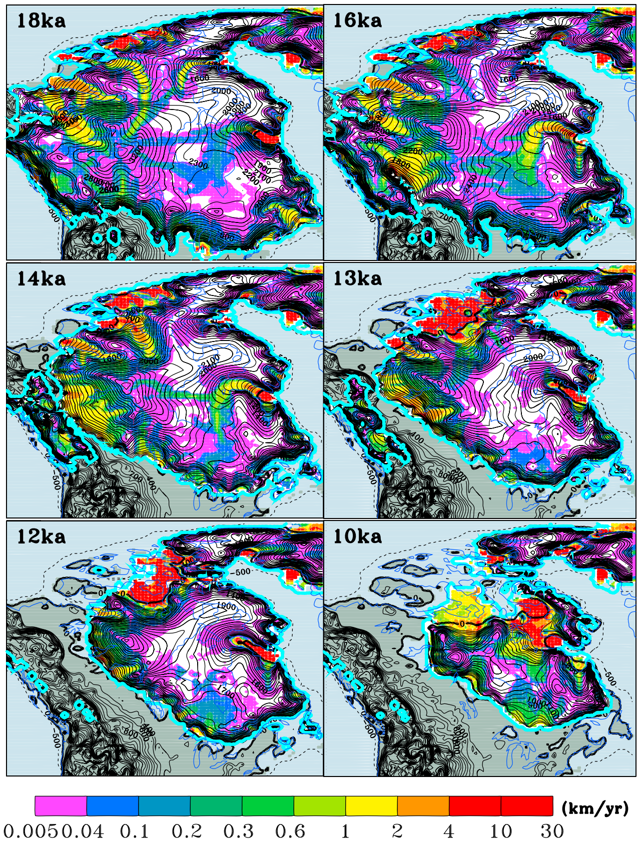


Figure 2. Modelled weighted ensemble average of basal velocity for the Laurentide Ice Sheet at 18, 16, 14, 13, 12, and 10 kyr. All ages are calendar yr BP.

[14] Further switches in streaming flow are seen at 13 kyr, with mass loss focussed in the marine channels and sounds in the Canadian Arctic and reduced streaming in the south-western sector. The M'Clure Strait ice stream (19) continues to expand southward whereas the neighbouring catchment of the Amundsen Gulf ice stream (18) is slightly smaller than at 14 kyr. The large tributary of the Hudson Strait ice stream (24) has disappeared. The streaming in the Lake Winnipeg/Red River (27, 28) area has decreased (shorter, less well-defined tributaries) but large areas of streaming flow have developed in northern Saskatchewan and northern Alberta near Lake Athabasca. No previous work has specifically invoked ice streaming in these locations. By 12 kyr, these ice streams are shorter and some have switched off altogether. Indeed, the most noticeable change at 12 kyr is the reduction in streaming activity, in part due to the retreat of ice sheet onto a predominantly hard bed, although ice streaming has switched on to the south-west of James Bay (26), where geological evidence has been reported. At 10 kyr, ice streams at the northern margin remain indistinct from a large ice shelf and the Hudson Strait ice stream terminus has retreated along the channel and now captures more ice from Foxe Basin. The location of streaming has switched positions again at the southern margin and is found north of Lake Winnipeg, where no geological evidence has been reported, although the switching near James Bay (26, 33) has been hypothesised. The ice stream depicted flowing northwards across Ungava Bay (17) at 10 kyr has also been reported as a short-lived readvance around this time.

4. Discussion and Conclusions

[15] The model successfully reproduces all of the large marine-terminating ice streams in topographic troughs previously hypothesised from geological evidence, supporting the notion that they were all important drainage conduits during full glacial conditions. The model also captures some, but not all, of the major terrestrial ice streams at the southern margin of the ice sheet that terminated on land but it is clear that terrestrial ice streams represent a more complex feature for the model to reproduce. Indeed, it fails to reproduce streaming velocities where evidence appears to be robust and when the ice sheet margin is lobate, presumably driven by higher velocities (e.g., the Des Moines and James at 18 and 16 kyr [see *Patterson*, 1998]). These cases are vital for evaluating the success of the model and guiding its future development. Indeed, terrestrial ice streams at the southern margin are thought to have been influenced by changes in basal hydrology, more akin to surging behaviour [see *Clayton et al.*, 1985], and it is anticipated that the planned incorporation of basal hydrology into the GSM will lead to improved results in this regard. In the context of the ongoing calibration of the GSM, persistent data-model misfits may also be used to further constrain the probability distribution for the derived deglacial ice sheet chronology.

[16] A further test of the model would be to examine whether it can reproduce ice streams at the right time (as well as the right place) but the scarcity of well-dated paleo-ice streams prevents a more rigorous analysis and there is a pressing need for further work in this regard. The data-model comparison does, however, provide several new insights regarding the temporal evolution of ice streaming during deglaciation. First, although some marine-based ice streams persist throughout deglaciation, where topography allows them to migrate up ice, the focus of mass loss associated with ice streaming clearly changes through time (Figure 2). The implication for contemporary ice sheet dynamics is that areas currently observed to be in balance may not be immune from dynamic changes over millennial time-scales. Moreover, both observations [*Anandakrishnan and Alley*, 1997] and modelling [*Payne*, 1999] of the West Antarctic Ice Sheet reveal fluctuations in ice stream behaviour over century-millennial timescales, even without major external forcing.

[17] Second, some marine-based ice streams migrate inland during ice margin retreat (e.g., M'Clure Strait); whereas others retreat to a point and then switch off (e.g., Laurentian Channel). This implies ice streams do not always persist during deglaciation and is probably related to conditions upstream of their onset zones inhibiting streaming flow (such as advection of cold-based ice) and/or as a result of ice or (potentially in more complete models) subglacial meltwater in their catchments being captured by neighbouring ice streams [*Payne and Dongelmans*, 1997; *Anandakrishnan and Alley*, 1997].

[18] Third, and related to the above, neighbouring ice streams can retreat in radically different styles, with one ice stream expanding, while the other remains unchanged or even shrinks. Contrast, for example, the up-ice expansion of

the M'Clure Strait ice stream (19) between 16 and 14 ka compared to the neighbouring ice stream in Amundsen Gulf (18), which maintains a similar size. This implies that neighbouring ice streams can behave quite differently in a deglaciating ice sheet [see *Stokes et al.*, 2009].

[19] Fourth, modelled ice streams exhibit quite different characteristics in their onset zones. The Hudson Strait ice stream (24) has an extensive tributary system at 18 kyr, whereas the Lancaster Sound ice stream (22) has a more limited tributary. Furthermore, onset zones vary over time, implying that their migration, and that of associated ice divides, is a key component of ice stream dynamics. Monitoring contemporary ice streams is often focussed on portions downstream but it would appear that dynamic changes also occur upstream and we note recent work in this regard in West Antarctica [*Conway and Rasmussen*, 2009].

[20] In conclusion, a data-calibrated GSM with thermo-mechanically-coupled shallow-ice physics can reproduce most geologically inferred ice-streams as well as indicate ice streams where geological evidence has not been widely reported but where it appears to exist. Further important results from this exercise are that (1) the focus of mass loss associated with ice streaming changes through time, (2) major changes in ice stream activity are to be expected in a deglaciating ice sheet, (3) most terrestrial ice streams do not persist during deglaciation, and (4) dynamic changes in ice stream onset zones and tributaries take place even in the absence of marked changes downstream. Subsequent to the inclusion of explicit basal hydrology and higher order ice dynamical stress components, future work will examine the spatial and temporal sensitivity of ice streams to uncertainties in both climate forcing and basal processes. This should lead to a refined understanding of the factors that are most important in influencing their behaviour (e.g., climate versus ice dynamical versus subglacial).

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